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The results will be compared with the predictions of the gravity wave model of Dewan (1990, 1991) and shown to be in reasonable agreement thus lending further support to the local wave-cascade hypothesis.

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MESOSPHERIC OH AIRGLOW TEMPERATURE FLUCTUATIONS:
A SPECTRAL ANALYSISE.M. Dewan¹, W. Pendleton², N. Grossbard³, and P. Espy²

Abstract. A field campaign under Project MAPSTAR was conducted in Colorado during May-July 1988. As part of this effort ground based measurements of OH airglow rotational temperatures at 85 km were made by means of an infrared Fourier spectrometer (IRFWI of USU). These measurements employed a least squares spectral fitting technique involving the 3-1 Meinel band. These data and their estimated PSD's will be presented. The latter will be interpreted by means of the relations between gravity wave temperature fluctuations and vertical displacement fluctuations given by Makhlof et al (1990).

The results will be compared with the predictions of the gravity wave model of Dewan (1990, 1991) and shown to be in reasonable agreement thus lending further support to the local wave-cascade hypothesis.

Introduction

A field campaign was conducted in Colorado during May through July 1988 under project MAPSTAR of AFOSR. Its purpose in part was to study the effects of gravity waves upon the temperature of the OH airglow layer located around 85 km altitude. To do this we employed the IRFWI Fourier spectrometer of Utah State University from which, by means of the 3-1 vibration transition rotational lines in the Meinel band (near 1.53μ), we obtained the OH rotational temperatures by employing a technique similar to that described in Hill et al. (1979). The instrument had a field of view of 0.8° and a resolution of 2.5 wave numbers. Due to the high throughput of the IRFWI instrument, temperatures of the OH Meinel band in question could be extracted at around a 30 second temporal resolution with a relative error of between $\pm 0.5^\circ \text{ K}$ and $\pm 1.5^\circ \text{ K}$ for each of the 23 nights of observation. For pointing angles both zenith and, more often, 25° elevation angles were employed.

The main purposes of the present paper are to exhibit the power spectral densities (PSD's) of these temperature fluctuations and to compare them with the theoretical expectations based on Dewan (1990, 1991).

Temperature Data and Spectral Analysis

Figure 1 shows a representative example of the temperatures over one of the 23 nights of observations. The obvious gaps, due to realignment procedures, were present on all nights and they made direct spectral analysis impossible.

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On average there were 4 such gaps per night constituting, on average, about 1/4 to 1/5 the duration of each night's observation interval. Another problem was that the sampling intervals varied somewhat over the night. This however was solved by cubic interpolation to create evenly spaced samples. Five nights had an average sampling rate of 19.9 sec. (which differed from night to night, $\sigma = 0.2$) and the average observation time was about 5 ± 1 hrs.; the remaining 18 nights had an average sampling interval of 39.5 sec. ($\pm .5$) and duration 4 ± 1.5 hrs.

Two independent methods were used to obtain the PSD's. The first was the well known one of Blackman and Tukey (1959), (BT), and in this case the problem of gaps was solved due to the fact that there were no gaps in the autocorrelations. Ninety percent of possible lags were used (instead of the usual 10%). The autocorrelations were obtained by the zero-padded FFT method of Oppenheim and Schaffer (1975) and we used the "biased" version (normalized by total number of data points) to avoid the subsequent occurrence of negative PSD's. Prior to autocorrelation the data were demeaned and prewhitened by $x'_i = x_i - \alpha x_{i-1}$. The autocorrelations were tapered by a triangular window (Bartlett) and the (whitened) PSD obtained via zero-padded FFT. "Post-darkening" was obtained by dividing the white spectrum by $(1 + \alpha^2 - 2\alpha \cos(\omega \Delta t))$ where Δt is the sampling interval (c.f. Blackman and Tukey, (1959)).

Straight lines were least square fitted to both the white and post darkened spectra. The values of $\alpha = 1.0, 0.95$, and 0.90 gave whitened slopes of $+0.3, +0.006$ and -0.3 respectively hence $\alpha = 0.95$ was chosen for the subsequent analysis. The fitted lines extended from a minimum frequency (with period equal to 1/3 the observation interval) to the maximum

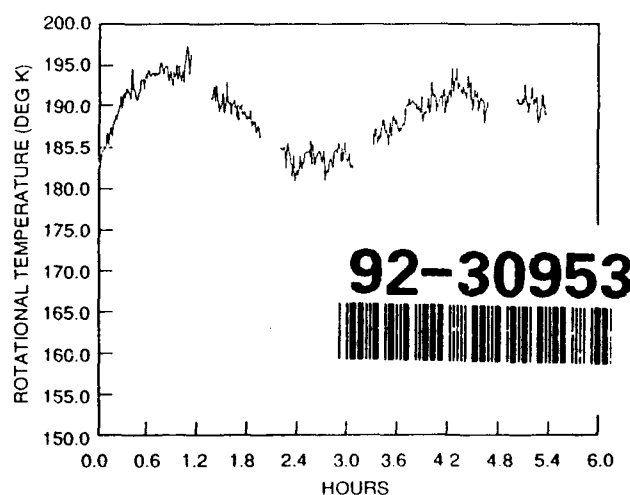


Figure 1. Rotational Temperatures (3-1 Vibrational Transition) Day 194, 1988.

frequency of 2.5×10^{-3} Hz which is approximately 80% of the Brunt Vaisalla frequency (i.e. maximum gravity wave frequency). This seemed to avoid noise, acoustic compression and other contaminants.

It should be mentioned that we have made the usual assumption (made in lidar and radar observations) that within such a frequency range the oscillations are due exclusively to gravity waves.

The other PSD technique used was the maximum entropy method (MEM) of Burg using the "Burg algorithm" (see Marple, 1987, p. 213). This was applied to each "piece" of data between the gaps which, in effect, supplied an alternate solution to the gap problem. We used 7 linear predictor coefficients the choice being based on the relatively slow decrease of prediction error, beyond that point.

As will be shown, the two methods gave very similar results and, in addition, they both gave appropriate responses to a family of known synthetic signals. Figure 2 shows a superposition of PSD's obtained by both methods for the data in Figure 1. In all 23 cases the MEM result looked like a smoothed version of the BT result. Straight lines were fit over the same ranges in the MEM PSD's as in the BT PSD's. Results from the two methods did not perfectly coincide on an individual basis, but there was very good overall statistical agreement as will be seen.

Theory for $\Psi_T(\omega)$, the PSD of Time Variations of Temperature

The following assumes that temperature variations between frequencies $\omega = \omega_i$ (inertial) and N (buoyancy frequency) are

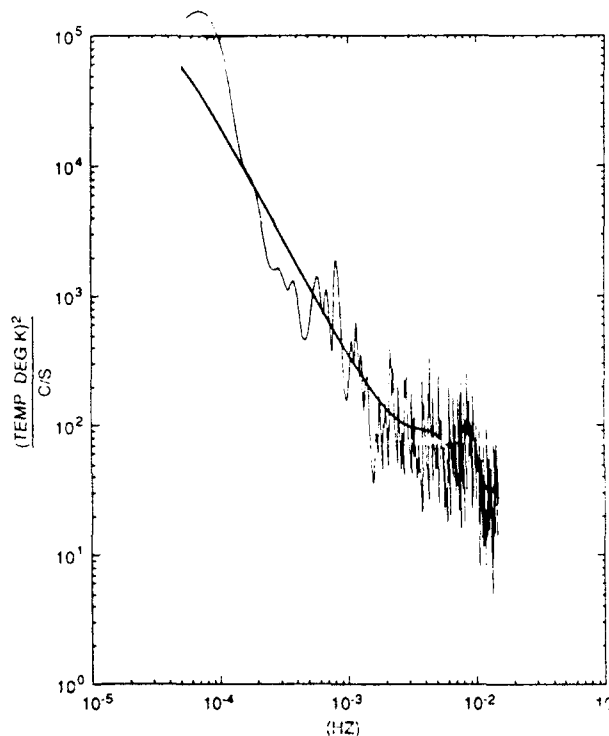


Figure 2. PSD's of Temperatures Given in Figure 1. Both Blackman and Tukey and Burg techniques (smoother one) were used. This shows an average case.

due to gravity waves (mean winds are ignored). See Appendix A for the justification. It was shown in Makhlof et al. (1990) that two separate physical mechanisms account for wave induced temperature variations. One of these is essentially acoustic in nature and the other is due to altitude variations of the fluid parcels. In the latter case the change in temperature, ΔT , due to a change in altitude, Δz , is given by

$$\frac{\Delta T}{T} = -\frac{(\Delta z)}{H} \left(\frac{\gamma - 1}{\gamma} \right) \quad (3.1)$$

where H is the scale height (which is 5.7 km at 85 km, U.S. Standard Atm., (1976)) and γ is 1.4 (specific heat ratio). For our purposes $\Delta z = 100$ m would give us approximately $\Delta T = 1^\circ\text{K}$. It can be shown that the above mentioned acoustic component can be ignored when $\omega < N/2$ at 85 km altitude. At higher frequencies the assumption needs further justification. We found $\omega < .8N$ to be satisfactory as was mentioned above.

How can the above be employed to derive $\Psi_T(\omega)$ from theory? In Dewan (1990, 1991) a similitude argument based on a wave cascade hypothesis led to the frequency PSD for vertical particle velocities due to gravity waves, $\Psi_v(\omega)$, given by

$$\Psi_v(\omega) = \left[\frac{\alpha' \epsilon}{\omega^2} \right] \left[\frac{\omega^2 - \omega_i^2}{N^2} \right] [2\pi] \quad (3.2)$$

in units of $(\text{m}^2/\text{s}^2)/(\text{Hz})$, where ϵ is the energy dissipation rate and α' is a constant of order unity. Term (1) is the wave energy density and is in some sense analogous to a similar relation given in Tennekes and Lumley (1972) in the context of turbulence. Term (2) is the fraction of wave energy in the form of vertical velocity variance at ω and is derivable from the polarization relations (cf. Garrett and Munk (1975) and Scheffler and Liu (1985)). The third term converts radians/sec to Hz. It can be shown that (using $\Delta z = 100$ m implies $1^\circ\text{K} = \Delta T$):

$$\Psi_{\Delta z}(\omega) = \frac{\Psi_v(\omega)}{\omega^2} = \Psi_T(\omega) \left(\frac{100\text{m}}{1^\circ\text{K}} \right)^2 \quad (3.3)$$

(compare Garrett and Munk (1975)), thus for $\omega \gg \omega_i$ we have

$$\Psi_T(\omega) = \frac{\alpha' 2\pi \epsilon}{N^2 \omega^2} \left(\frac{1^\circ\text{K}}{100\text{m}} \right)^2 \quad (3.4)$$

Appropriate values for the parameters are $N = 0.02$ rad/sec (Dewan and Good (1986)), and $\alpha' = 2.2$ (Dewan, 1990). Eq. (3.4) thus relates ϵ to measured values of $\Psi_T(\omega)$ and vice versa. For this reason, if the above is valid, it may be possible to estimate average ϵ over certain altitude regions by means of interferometric measurements of airglow temperature fluctuations.

tuations (e.g. OH at 85 km, Na at 90 km, and OI at 95 km). Equally promising are the new lidar temperature measurement techniques described by Von Zahn (1987), Gardner (1991), and She (1991). In the meantime it is essential to test the theory by experimentally measuring ϵ at the same time that $\Psi_T(\omega)$ is obtained. One way to do this is by means of Doppler broadening of radar returns as described in Hocking (1988). Radar measurements of $\Psi_v(\omega)$ would be very useful in addition of course.

Comparison Between Theory and Experiment

Figure 3 summarizes all the fitted lines (23) for the BT-method derived $\Psi_T(\omega)$ spectra. The average slope was -1.7 ± 0.3 . The MEM spectra, while very smooth, gave similar line fits with slopes -1.7 ± 0.2 on average. These results are consistent with the theoretical value of -2.0 . (On the other hand if a larger data set were to narrow the spread around -1.7 , the proverbial $-5/3$ slope would be a more valid choice (cf. the appendix of Dewan, 1990)). We now compare $\Psi_T(\omega)$ amplitude with the theoretical connection with observed ϵ values.

Solving Eq. (3.4) for ϵ ,

$$\begin{aligned}\epsilon &= \Psi_T(\omega) \omega^2 \cdot \frac{N^2 (100)^2}{2\pi \alpha'} \\ &= (0.29) [\Psi_T(\omega) \cdot \omega^2]\end{aligned}\quad (4.1)$$

Figure 4 shows the 23 fitted lines of $\Psi_T(\omega) \cdot \omega^2$. As can be seen they fall mainly in the amplitude range between 10^{-2} and 10^{-1} which, from Eq. (4.1) gives the approximate range ($0.003 < \epsilon < 0.03 \text{ m}^2/\text{s}^3$). This overlaps some observations of ϵ (at another time and geographic location) by Hocking (1988) in the altitude range of 80-90 km. These results gave ($0.01 < \epsilon < 0.1 \text{ m}^2/\text{s}^3$). This outcome is sufficiently encouraging to

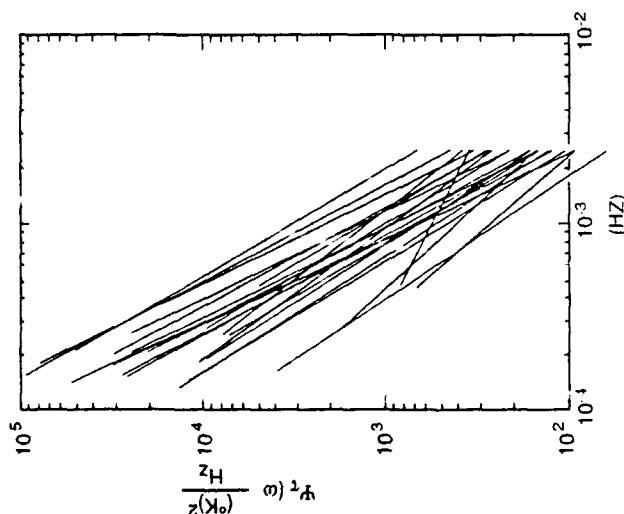


Figure 3. Summary of all 23 PSD's. Fitted lines from BT derived PSD's used here.

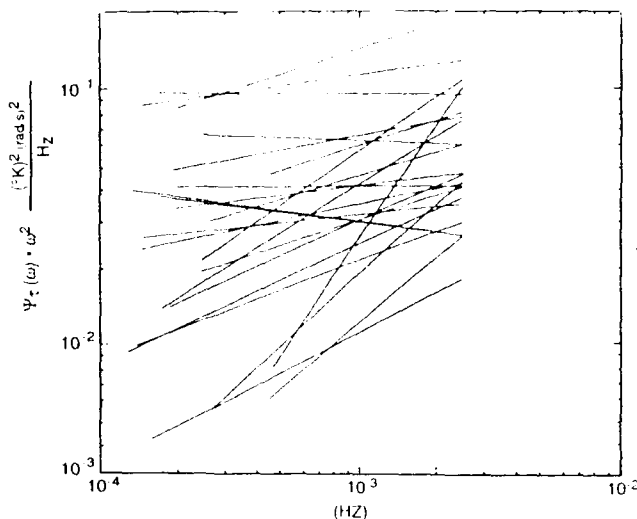


Figure 4. Summary of BT Fitted Lines for $\Psi_T(\omega) \cdot \omega^2$.

suggest that it would be most desirable to conduct a direct test by means of simultaneous radar (ϵ) and interferometer (or lidar) (T) measurements.

A word of caution is needed. Under certain circumstances the measured $\Psi_v(\omega)$ is not flat (c.f. Eq. 3.2) but instead can have a negative slope (Ecklund et al. (1986) and Sato (1990)). Why $\Psi_v(\omega)$ occasionally has a negative slope ($-5/3$) has not yet been fully explained. Since Eq. (3.2) holds only in a fluid at rest (i.e. intrinsic coordinates) David Fritts (in private conversation) has suggested that Doppler effects may be the cause of the observed negative slopes. Needless to say the results of the present paper would not be valid under those conditions where Eq. (3.2) is not valid. See also Van Zandt et al (1991).

Conclusions

OH rotational temperatures were measured (for 23 nights) and the $\Psi_T(\omega)$ PSD's obtained. The results appear to agree with the theory of internal gravity wave PSD's presented in Dewan (1990, 1991) and deduced values of ϵ overlap the range experimentally measured. Direct radar and interferometer tests were suggested and if they result in confirmation of the theory, one will have further evidence that the wave cascade idea first suggested in Dewan (1979) is valid. We would also have a new method to obtain ϵ via interferometer or lidar temperature measurements (under appropriate conditions).

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Appendix A

On the Demonstration That Mean Winds at OH Altitudes Have Negligible Effects on our $\Psi_T(\omega)$ Measurements.

During the 1988 Colorado campaign Prof. S. Avery of the University of Colorado made meteor wind measurements from the Platteville radar site which was centrally located with respect to our instruments. The device, known as MEDAC (Wang et al, 1987), was used. In this way we obtained a mean wind altitude profile for days 127 to 142, and another average for 199 to 205. These intervals covered 16 of the 23 nights of observation. At no time did these mean winds, u , exceed 15 m/s in the important OH emitting range of 83-87 km. We now assume this to be at least approximately valid for the entire campaign.

According to Van Zandt et al (1991) the parameter $u_* \equiv u/c_*$ (where c_* is the characteristic horizontal phase trace velocity) determines the doppler effect on temporal spectra.

Using $c_* = \frac{N}{k_*}$ and $k_* = \frac{2\pi}{20 \times 10^3 \text{ m}}$ from Smith et al (1987)

one obtains $u_* < 0.25$. Figure 8 of Van Zandt et al (1991) shows that, in this case, there would be negligible effects on our spectral measurements.

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